# From arid to humid – The Jurassic–Cretaceous boundary interval in northern Germany

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A B S T R A C T

The Jurassic–Cretaceous boundary interval in northwest Europe is characterised by a distinctive climatic change from pronounced aridity toward more humid conditions. In order to better understand the timing and evolution of the environmental dynamics related to this climatic change, terrestrial and aquatic palynomorphs (spores, pollen, dinoflagellate cysts, freshwater algae) have been studied from two recently drilled cores. The cores, which are both located in the Lower Saxony Basin (northern Germany), provide two 139 m and 134 m thick non-marine successions embedded in a high-resolution biostratigraphic framework. The lower part of the cores can be attributed to the Münder Formation (upper Tithonian–middle Berriasian), the upper part to the Bückeberg Group (middle–upper Berriasian). For reconstructing the arid to humid climate transition 110 samples have been analysed for their palynological content.

In the Lower Saxony Basin, the late Jurassic–earliest Berriasian vegetation was dominated by cheirolepidiacean conifer forests growing under arid conditions along a coastal belt. A marked decline in cheirolepidiacean pollen and the spread of pioneering plants records an increase in seasonal humidity in the early–mid Berriasian. In the mid–late Berriasian the presence of highly diverse floras reflects even more humid and warm conditions. Fluvio-deltaic mixed swamp forests, which grew landwards behind the coastal belt were composed of conifers, ferns, lycopods, horsetails, bryophytes, ginkgos, cycads and Bennettitales. Cheirolepidiaceans, probably forming part of mangrove-type plant communities, remained an important component of the coastal flora. The climatic shift correlates to the upper *Subcraspedites lamplughi*–middle *Heteroceras kochi* ammonite zones of the marine Boreal zonation scheme and is synchronous to the early–mid Berriasian shift from arid to humid recorded from southern England.

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## 1. Introduction

The Jurassic–Cretaceous (J–K) boundary beds of northwest Europe (southern England, northern Germany, northern France) show multiple evidence for a pronounced climatic change (Hallam, 1984). Widespread arid conditions prevailing in the late Kimmeridgian–Tithonian were replaced by a more humid climate in the mid Berriasian. This change in climate is indicated by several lines of evidence including shifts of the sediment patterns. Thick evaporite-dominated sequences (gypsum, halite) are typical for the latest Jurassic, several coal seams are well documented from the earliest Cretaceous (references in Pelzer and Wilde, 1987; Batten, 2002). Further support comes from clay minerals

(e.g., Sladen, 1983; Hallam, 1984; Allen, 1998; Schnyder et al., 2006) and spore-pollen data (e.g., Burger, 1966; Vakhrameev, 1981; Sladen and Batten, 1984; Hunt, 1985; Heunisch and Luppold, 2015). Palaeobotanical findings suggest that the climatic conditions of the late Kimmeridgian–Tithonian were similar to those of the present-day Mediterranean, with arid, hot summers and wet, cool winters (e.g., Francis, 1984; Allen, 1998; Abbink et al., 2001).

A vegetation change from draught-adapted Cheirolepidiaceae-dominated conifer forests to mixed forests with diverse fern associations is recorded for the earliest Cretaceous. This shift toward more humid conditions has been documented not only in Europe (e.g., Döring, 1965; Burger, 1966; Mamczar, 1968; Norris, 1969; Hunt, 1985) but also in northwest China (Zhang et al., 2014).

During the Late Jurassic–Early Cretaceous, the flora of the Northern Hemisphere can be attributed to two different palaeobiogeographic areas, the Arctic Subprovince (Siberian-Canadian) of the temperate high latitudes and the European-Sinian Subprovince (Vakhrameev, 1978; Batten, 1984; Ziegler et al., 1993; Herngreen et al., 1996). The

latter encompasses the tropical and subtropical low and middle latitudes and is characterised by a Late Jurassic arid climate. The floras of the Siberian-Canadian Subprovince were less diverse than those of the European-Sinian Subprovince (Vakhrameev, 1991; references in Pott et al., 2014). Palynological data from northern Canada document an increase in humidity not earlier than late Valanginian or early Hauterivian, thereby post-dating the change observed in the European-Sinian Subprovince by ca. 8 million years (Galloway et al., 2013).

It is still a matter of debate to which extent the J–K climatic shift was linked to the breakup and subsequent fragmentation of Pangea and to changes in monsoonal patterns (Parrish, 1993; Ziegler et al., 1993; Bush, 1997). According to Abbink et al. (2001) the observed change in climate reflects a regional signal limited to northwest Europe. Sladen and Batten (1984) associated it with the uplift of the Anglo-Brabant Massif. Using the General Circulation Model from Valdes et al. (1996) the palaeoenvironmental interpretations of Allen (1998) attributed the J–K shift to the widening of the evolving Proto-Atlantic. According to the latter author, the large-scale tectonic reconfiguration resulted in the permanent establishment of year-round westerlies in the Berriasian, increasing thereby the total precipitation rate during Cretaceous winters.

For northern Germany the climatic conditions in the Tithonian and Berriasian have been described earlier based on clay minerals, sedimentology and palynology (Pelzer and Wilde, 1987; Gramann et al., 1997), but so far no biostratigraphically calibrated data across the J–K boundary interval are available. Quantitative sporomorph data have been published from the Wealden strata of the eastern Netherlands (Burger, 1966), but not from northern Germany. The biostratigraphic control of Burger (1966) remains vague, since no ostracod biostratigraphy is available.

In the current study new quantitative palynological findings from two cores of the Lower Saxony Basin (LSB) are used to reconstruct the environmental and climatic changes across the J–K boundary interval. The records are integrated into the existing biostratigraphic framework of the non-marine J–K successions in northwest Europe based on ostracods (e.g., Wolburg, 1949, 1959; Bischoff and Wolburg, 1963; Anderson and Hughes, 1964; Kemper, 1973; Elstner and Mutterlose, 1996) and palynomorphs (e.g., Dörhöfer, 1977; Dörhöfer and Norris, 1977; Wimbledon and Hunt, 1983; Hunt, 1985, 2004; Abbink et al., 2001; Lindström and Erlström, 2011; Schneider et al., 2018).

## 2. Geological setting

In Late Jurassic–earliest Cretaceous times, northwest Europe was subdivided into several small landmasses and isolated epicontinental basins (Ziegler, 1982, 1990) linked to a global sea-level lowstand (Fig. 1; Haq et al., 1988; Sneider et al., 1995; Haq, 2014). Non-marine sediments of the Purbeck and Wealden facies accumulated in southern England, the Paris Basin and northern Germany. The type locality of the Purbeck is located in Dorset, southern England, where the Purbeck Limestone Group rests unconformably on marine strata of the late Tithonian Portland Group. The Purbeckian strata are overlain by the Wealden Group (Fig. 2). The term "Wealden" was originally used for Lower Cretaceous non-marine deposits of southern England, but is nowadays also applied to non-marine strata in other areas, such as the LSB. In England, Wealden-type sediments range from the latest Berriasian–early Aptian. The German Wealden facies in contrast only covers the mid–late Berriasian.

In northern Germany, non-marine Purbeck- and Wealden-type sediments were deposited in two main subbasins, the LSB and the Südmecklenburg-Brandenburg Basin (SM-BB; Fig. 1). To the south, the LSB was bordered by the Rheno-Bohemian landmass and to the north by the Pompeckj Swell (e.g., Kockel et al., 1994; Gramann et al., 1997). The Purbeck-type deposits investigated here belong to the middle Münder Formation (OM 4 = Ober Malm 4) of latest Tithonian age and the upper Münder Formation (OM 5–6 = Ober Malm 5–6) of early Berriasian age. The units OM 5 and OM 6 were formerly known as the Katzberg and Serpulit members. The Münder Formation comprises
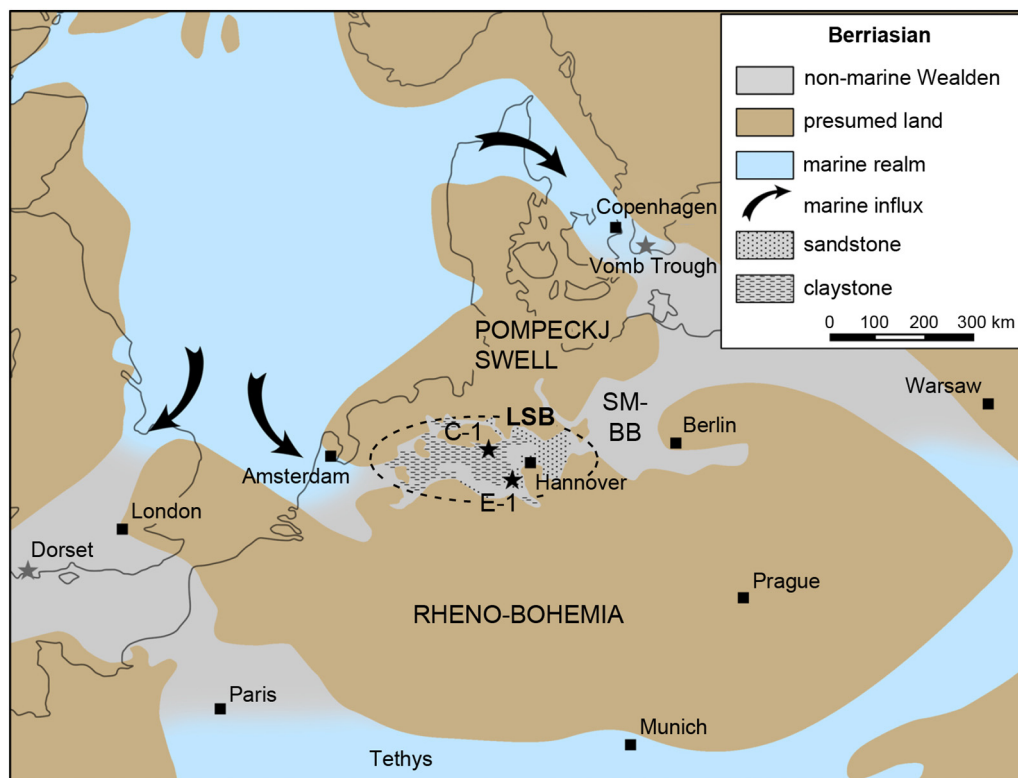


**Fig. 1.** Palaeogeographic map of northwest Europe for the Berriasian showing the positions of cores E-1 and C-1 in the Lower Saxony Basin (LSB), the Purbeck type section in Dorset and the Vomb Trough in Sweden. Modified after Schott et al. (1967/69), Ziegler (1982, 1990), and Mutterlose (1992). SM-BB: Südmecklenburg-Brandenburg Basin.
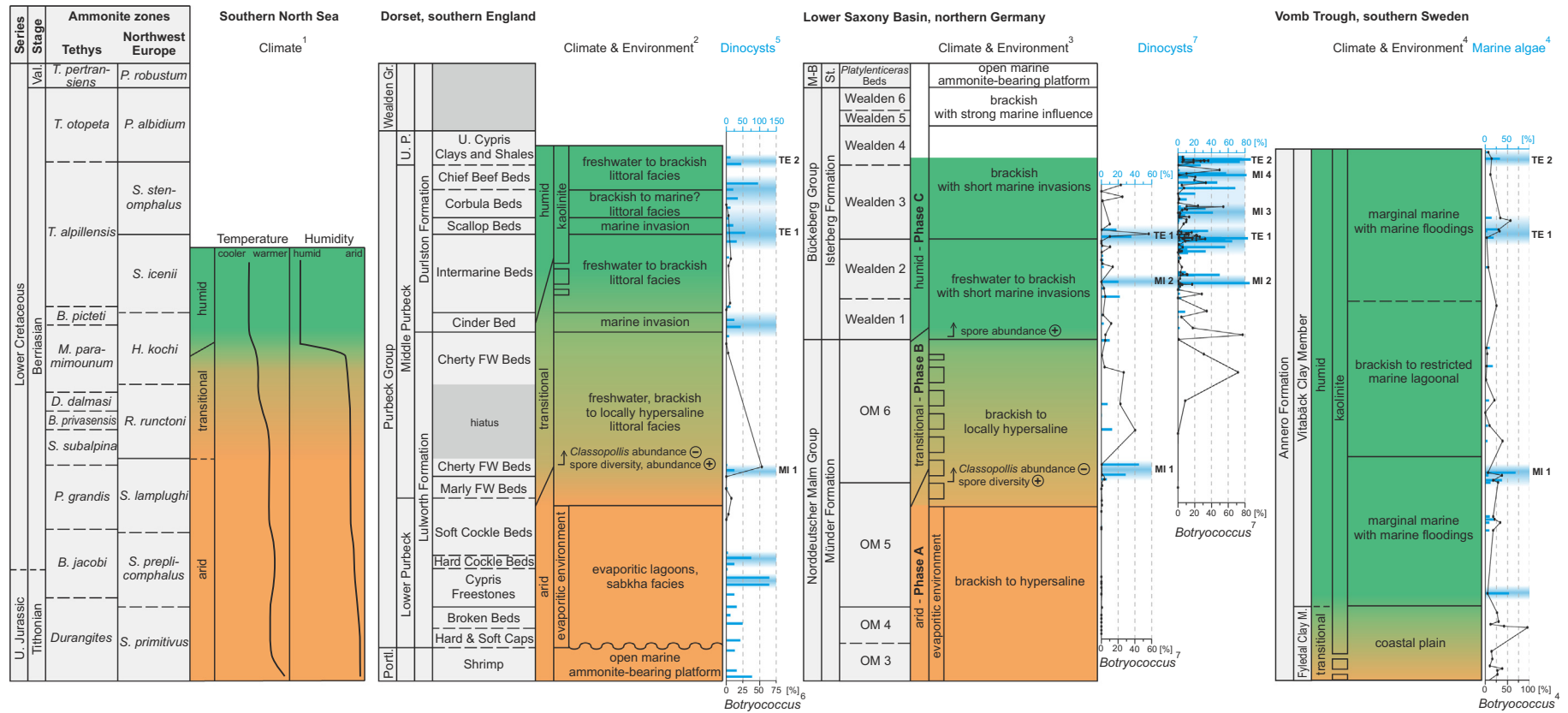
**Fig. 2.** Palaeoclimatic and palaeoenvironmental changes in northwest Europe across the J–K boundary interval. The palaeoclimatic reconstructions show a shift from arid conditions to a more humid, warm climate based on data from the marine Boreal of the southern North Sea (Abbink et al., 2001) (1), the non-marine Purbeck (Schnyder et al., 2006, 2009) (2), the LSB (this study) (3) and the Vomb Trough (Lindström and Erlström, 2011) (4). Purbeckian miospore and microplankton data are from Hunt (1985, 2004) (5) and Schnyder et al. (2009) (6); lithostratigraphic data are adopted from Wimbledon and Hunt (1983), Clements (1993) and Westhead and Mather (1996). The stratigraphic correlation, TE's (transgressive events) and MI's (marine incursions), and microplankton data for the LSB are from Schneider et al. (2018) (7). *Calpionella alpina* has been selected by the International Commission on Stratigraphy as the primary marker for the base of the Berriasian in the Tethys (e.g., Wimbledon et al., 2011). Caution is necessary, for no ammonite scheme in the Tethys or any part of the boreal near the J–K boundary is calibrated (e.g., Schnabl et al., 2015). In Durlston Bay, Dorset, the J–K boundary may be placed in the Cockle Beds.
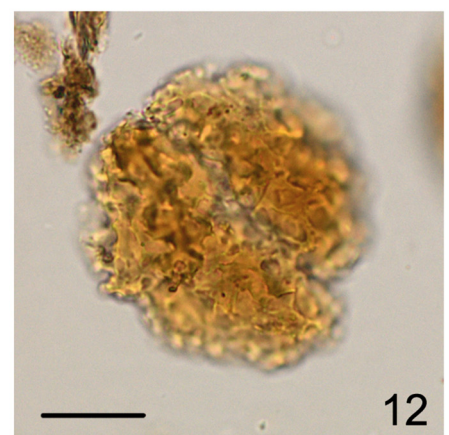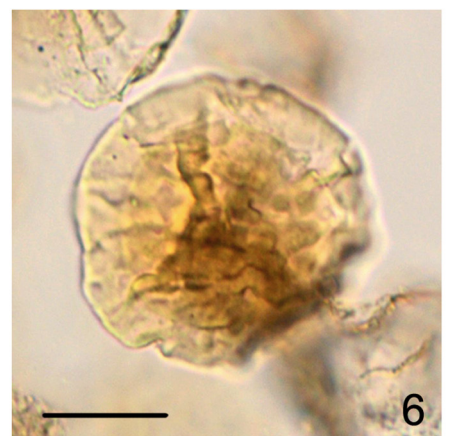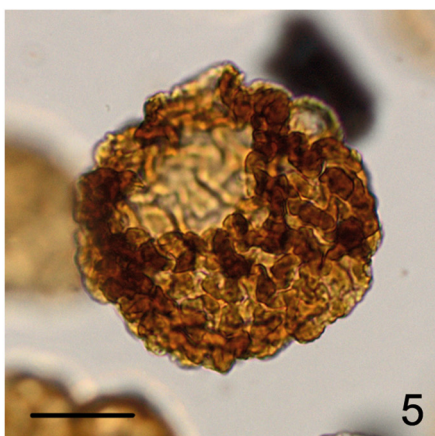
marlstone-gypsum alternations including oolites, serpulites and stromatolitic limestones. It is overlain by the non-marine Bückeberg Group or Isterberg Formation (Wealden 1–6; Erbacher et al., 2014). These Wealden-type deposits accumulated under fluctuating freshwater to brackish-marine conditions. Seven short-lived marine flooding events, documented by dinoflagellate cysts and foraminifera, can be used as marker horizons for intra- and interbasinal correlation (Schneider et al., 2018). In the central part of the LSB the Wealden-type deposits are represented by TOC-rich claystones (up to 18% TOC; Rippen et al., 2013) reaching a maximum thickness of 700 m (Schott et al. 1967/69; Mutterlose and Bornemann, 2000). Along the northern and southern margins of the basin, sandstones, coal seams and *Neomiodon* (bivalve) limestones occur. Along its southeastern margin, in the Deister-Hils Embayment, a transition from fluvial deposits of an upper deltaic plain to lower deltaic and lagoonal clastics occurs. All these environments are associated with coal seams (Pelzer, 1998; Pelzer et al., 1992).

## 3. Material and methods

The Eulenflucht-1 well (E-1) was drilled in 2009 in the Deister-Hils Embayment. The 328 m deep well cored the lower part of the German Wealden and almost the entire Upper Jurassic. For this study, a 139 m thick interval encompassing the J–K boundary interval has been investigated. From the second core (C-1), which was drilled by ExxonMobil at the northeastern margin of the basin, a 134 m thick interval has been investigated.

### 3.1. Palynology

Sample preparation was done at the BGR (Bundesanstalt für Geowissenschaften und Rohstoffe) following standard palynological techniques using hydrochloric and hydrofluoric acids. The material was sieved with a mesh size of 10 μm before mounting on specimen slides (Gray, 1965; Wood et al., 1996). Terrestrial spore-pollen assemblages were analysed from a total of 110 samples and a minimum of 200 sporomorphs was counted (61 samples from E-1 and 49 samples from C-1). From the E-1 core 19 out of 61 samples were barren or nearly barren of sporomorphs. In core C-1 all 49 samples yielded sporomorphs. Relative abundances were calculated and plotted using the Tilia and TgView programmes (Grimm, 1991–2001). Sporomorph assemblages were established by constrained cluster analysis using CONISS (Grimm, 1987) within Tilia. Slides are stored in the collection of the BGR, Hannover. Sporomorphs were identified based on Döring (1965), Burger (1966), Norris (1969), Dörhöfer (1977), Boulter and Windle (1993), Schrank (2010), and Lindström and Erlström (2011).

## 4. Spore and pollen abundance and distribution

A total of 21 spores and 10 pollen taxa were identified on the genus and species level, other sporomorphs were treated as groups (e.g., bisaccate pollen). Selected spores, pollen, and phytoplankton taxa are documented in Plate I. Preservation of palynomorphs in both cores is moderate, showing signs of post depositional degradation. A high degree of fragmentation hampers a clear differentiation of *Classopollis* pollen and small inaperturates like *Perinopollenites*, resulting in considerable amounts of taxonomically undifferentiated pollen. Thermal alteration of the organic matter is indicated by the orange brown colour of the palynomorphs (Plate I; thermal alteration index 3 after Staplin, 1969). The quantitative distribution patterns for both cores

are illustrated in Figs. 3 and 4. *Araucariacites* and *Callialasporites* were treated as one group because they are known to show polymorphism; transitional forms seem to exist between these two genera (Balme, 1995; Schrank, 2010). In both cores, the sporomorph assemblages are dominated by *Classopollis* spp., *Perinopollenites elatoides* and bisaccate pollen. The *Deltoidospora/Cyathidites/Dictyophyllidites* group (= *Deltoidospora*-type) constitutes nearly half of the spores recovered. Based on the sporomorph frequency distribution, four successive assemblages have been recognized for each of the two cores using constrained cluster analysis. The assemblages of core E-1 have been labelled as EA-1 to EA-4, those of core C-1 as CA-1 to CA-4.

### 4.1. Sporomorph assemblages of core E-1

The lowermost sporomorph assemblage EA-1 (142.0–128.0 m) is characterised by low abundance variations with *Classopollis* (avg. 59%) reaching peaks up to 80%. Spores are virtually absent (avg. 2%), bisaccate pollen account for avg. 6%.

In EA-2 (128.0–113.12 m) 13 out of 17 samples are barren or nearly barren of sporomorphs. The floras are characterised by a decrease in *Classopollis* abundance (avg. 32%) and an increase in bisaccate pollen (avg. 17%). *Cerebropollenites* pollen show peak values up to 13%, whilst spore abundances remain very low (avg. 3%). The diversity of spores increases toward the top of EA-2. Additionally, *Vitreisporites pallidus* pollen become a regular, accessory floral element henceforth.

EA-3 (113.12–67.0 m) shows significantly higher spore (avg. 20%) and lower pollen abundances than EA-1 and EA-2. Spores increase up to 67% in the middle part of the succession. Bisaccate pollen are relatively stable (avg. 14%) with only the lowermost sample showing high abundances (77%). *Classopollis* (avg. 23%) and *Perinopollenites* (avg. 21%) are the dominant pollen.

In EA-4 (67.0–5.90 m) spores are dominant (avg. 61%) and highly diverse. Most prominent are spores of the *Deltoidospora/Cyathidites/Dictyophyllidites* group (avg. 30%) and *Cicatricosisporites* (avg. 13%). In contrast, bisaccate pollen (up to 10%), *Classopollis* (avg. 16%), and *Perinopollenites* (avg. 10%) are rarer.

### 4.2. Sporomorph assemblages of core C-1

The spore-pollen assemblage CA-1 (287.35–269.0 m) is mainly composed of *Perinopollenites* (avg. 31%), *Classopollis* (avg. 25%), and bisaccate pollen (avg. 21%). *Cerebropollenites* is common (up to 15%), whilst spores are very rare (avg. 3%).

CA-2 (269.0–226.0 m) is marked by an increase in spore abundance (avg. 11%) and diversity. Bisaccate pollen (avg. 22%) remain stable, whereas *Classopollis* (avg. 20%) and *Perinopollenites* (avg. 25%) numbers decrease slightly.

CA-3 (226.0–221.0 m) was closely sampled (9 samples) to assess the variability of sporomorphs in an interval which includes an intercalated siltstone horizon. The assemblages show distinctive fluctuations, they are characterised by a dominance of spores (avg. 55%), with *Deltoidospora*-type (up to 50%) and *Gleicheniidites* (up to 11%) being very common. *Perinopollenites* (avg. 17%), bisaccate pollen (avg. 12%), and *Classopollis* (avg. 11%) are relatively uncommon.

The sporomorph frequency distribution of CA-4 (221.0–158.3 m) is similar to CA-2. It is characterised by moderate spore abundances (avg. 15%) which decrease up-section. Bisaccate pollen remain stable (avg. 15%) and *Classopollis* occurs in moderate abundance (avg. 15%) with peak values (up to 60%) in two samples in the upper part.

**Plate I.** Photomicrographs of selected spores and pollen grains and phytoplankton. Scale bar is 20 μm. (1) *Classopollis torosus* Burger 1965, tetrad (E-1, 129.16 m); (2) *Perinopollenites elatoides* Couper 1958 (C-1, 281.50 m); (3) *Perinopollenites elatoides* (E-1, 135.87 m); (4) *Classopollis torosus*, proximal pole (C-1, 227.70 m); (5) *Cerebropollenites macroverrucosus* (Thiergart 1949) Schulz 1967 (E-1, 129.16 m); (6) *Callialasporites dampieri* (Balme 1957) Dev 1961 (C-1, 262.55 m); (7) *Parvisaccites radiatus* Couper 1958 (E-1, 129.16 m); (8) *Vitreisporites pallidus* (Reissinger 1950) Nilsson 1958 (E-1, 55.45 m); (9) *Cicatricosisporites* cf. *australiensis* Potonié 1956 (C-1, 222.70 m); (10) *Deltoidospora*-type spore (C-1, 222.70 m); (11) *Celyphus rallus* Batten 1973 (C-1, 213.48 m); (12) *Botryococcus* sp. (E-1, 107.0 m).
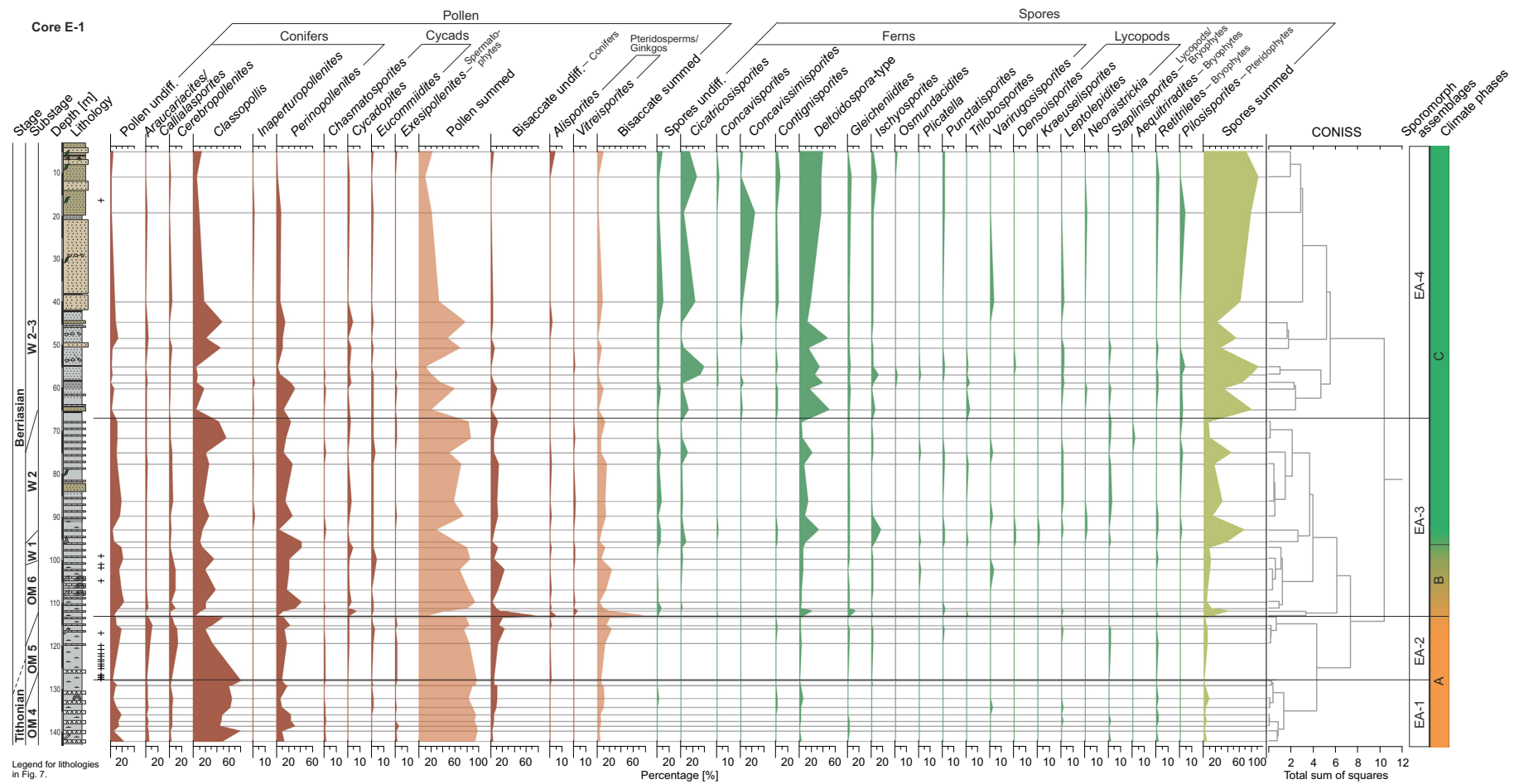
**Fig. 3.** Relative spore and pollen abundances (%) in core E-1 plotted against stratigraphy, depth (m) and lithology. The sporomorph assemblages are based on CONISS cluster analysis. Biostratigraphy is following Elstner and Mutterlose (1996). Samples which are barren of palynomorphs are marked by crossed bars. Climate phases are referring to humidity with orange colour representing low, green colour high moisture availability.

*Perinopollenites* follows an opposite trend and decreases in the upper part (avg. 17%).

## 5. Vegetation reconstruction

The palynofloras of the J–K boundary interval in the LSB are composed of sporomorphs derived from various plant groups. Information on the botanical affinities of the quantitatively important spore and pollen genera is summarized in Table 1. According to this compilation, seven principal groups can be differentiated: cheirolepidiacean conifers, other conifers, cycads, pteridosperms/ginkgos, ferns, lycopods, and bryophytes. The extinct conifer family Cheirolepidiaceae is highlighted here due to its significance for climatic interpretations.

### 5.1. Sorting effects

The distribution of spores and pollen in sediments, primarily determined by the vegetation, is also influenced by sorting effects during transportation and deposition (e.g., Chaloner and Muir, 1968; Batten, 1974). This is reflected by average spore (non-saccate pollen) abundances ranging between 14% (65%) in pure claystones and 36% (46%) in siltstones (Fig. 5) in the lower–middle part of core C-1. A similar spore-pollen distribution and relatively stable numbers of bisaccates (16–22%) have also been observed in the Wealden of south-east England by Batten (1974). The negative correlation of spores and pollen can be explained by the greater specific density of spores which therefore tend to settle in settings closer to the source. Pollen in contrast dominate in more distal positions, where dinoflagellate cysts are more common. Due to this different hydrodynamic behaviour of spores and pollen, the more distal core C-1 yields significantly less spores than core E-1, although palynological data from both cores reflect the same climatic trend (Fig. 6). The high amounts of spores and the absence of dinoflagellate cysts in the upper part of core E-1 is explained by the transition toward a lower deltaic setting with phytoplankton assemblages being exclusively composed of *Botryococcus*.

In Fig. 5 bisaccate conifer pollen show no correlation with lithology. The distal core C-1, however, yields higher numbers of bisaccates than core E-1, which is positioned more proximal in the Deister-Hils Embayment. This can be explained by the high buoyancy of bisaccate pollen, their air sacci allow for a longer transport and thus enrichment in distal deposits (e.g., Chaloner and Muir, 1968). Further, bisaccates are more common in the fluvial and lacustrine deposits located further southeast in the upper deltaic plain (Pelzer, 1998) than in the lower deltaic plain deposits of core E-1. In this part of the basin westward transportation of bisaccate pollen may have been hampered by the presence of year-round westerlies, according to the palaeoenvironmental model of Allen (1998).

The sporomorph assemblages of the early Berriasian from both cores evidence a fundamental change of the vegetation, which cannot be explained by sorting effects and sea-level changes alone. The different composition of the spore-pollen assemblages in the OM 4–6 (avg. 3% spores in core C-1) and the Wealden 1 claystones (avg. 14% spores in core C-1) provides evidence for a distinctive climatic shift in the early–mid Berriasian. Supported by sedimentological evidence (see further down) the studied interval can be subdivided into three successive climate phases.

### 5.2. Climate Phase A (OM 4–5)

The palynoflora of climate Phase A (Fig. 6) corresponds to a typical Tithonian LSB assemblage as documented earlier by Heunisch and Luppold (2015). Phase A, represented by sporomorph assemblages EA-1 and EA-2 in core E-1, is characterised by a stable conifer association dominated by cheirolepidiaceans (up to 80% *Classopollis* pollen). Mesozoic *Classopollis*-producing conifers of the extinct family Cheirolepidiaceae thrived under warm, seasonal arid conditions

(e.g., Vakhrameev, 1981, 1991; Francis, 1983, 1984; Abbink et al., 2001). Cheirolepidiaceans are considered thermophilous and drought-resistant and were adapted to a wide range of habitats, ranging from well-drained upland (Vakhrameev, 1970, 1982) to coastal belt habitats (Hughes and Moody-Stuart, 1967; Batten, 1974; Watson, 1988; Heimhofer et al., 2008), even including the margins of hypersaline lagoons (Francis, 1984).

Bryophytes and pteridophytes are almost absent (avg. 2% spores) during Phase A. Most of the spores encountered in the two cores can be assigned to fern plants. The majority of Mesozoic ferns flourished under moist, rather warm conditions either in marshes, along river banks or as forest understorey (van Konijnenburg-van Cittert, 2002). Ferns require moisture for reproduction, high fern abundances and diversities are therefore indicative of humid conditions preferably in warm environments (e.g., Hallam, 1984; Abbink et al., 2004).

To obtain information on relative changes in humidity, the different plants discussed here have been grouped into xerophytes and hygrophytes based on their ecological preferences following the concept of earlier studies (Visscher and van der Zwan, 1981; Heimhofer et al., 2012). Following this approach, the xerophyte group is represented solely by *Classopollis*-producing conifers, whereas all spore plants are regarded as hygrophytic. Phase A shows constantly high xerophyte/hygrophyte ratios (avg. 94% xerophytes) indicating very low moisture availability. Sedimentological data provide additional evidence for arid conditions, namely the occurrence of gypsum layers which are accompanied by stromatolites and serpulids (Fig. 6). Arp et al. (2008) reported initial gypsum impregnation of stromatolite-forming biofilms in a hypersaline lagoon from the OM 4–5 of the Deister-Hils Embayment.

Another important plant group of Phase A are taxodiacean conifers, represented mainly by *Perinopollenites* pollen. Some authors suggest that *Perinopollenites* pollen were produced in coastal areas, under warm, dry conditions (e.g., van Amerom et al., 1976; Vajda, 2001). Others interpret high abundances of *Perinopollenites* as an indicator for a cooler, more humid climate (Abbink et al., 2004; Heimhofer et al., 2012). Since the fluvial facies of the Deister-Hils Embayment yields only small amounts of *Perinopollenites* (Pelzer, 1998), a coastal habitat is proposed for these taxodiacean conifers.

Araucariacean and cypressacean conifers, represented by *Araucariacites*/*Callialasporites* and *Inaperturopollenites* pollen (van Konijnenburg-van Cittert, 1971; Boulter and Windle, 1993; Balme, 1995), formed only a subordinate element of the vegetation. Araucariaceans often grow in coastal forests; dense populations indicate arid conditions. The extant genus *Araucaria* has a wide tolerance, ranging from rainforests to cool temperate forests (Harris, 1979; Reyre, 1980; Vakhrameev, 1991; Schrank, 2010).

In summary, the vegetation of Phase A consists of coastal conifer forests (cheirolepidiaceans, taxodiaceans, araucariaceans, cypressaceans) thriving under seasonally arid conditions, whilst pteridophytes and bryophytes are extremely rare.

### 5.3. Climate Phase B (OM 6)

Climate Phase B includes sporomorph assemblages from the lower part of EA-3 and from CA-1 (OM 6). In this interval a marked decline in cheirolepidiacean conifers is observed, interpreted to reflect an increase in humidity. This trend is accompanied by a spread of spore plant communities near the base of Phase B. *Vitreisporites*-producing pteridosperms, which are typically associated with warm and rather wet habitats (Pelzer, 1984; van der Burgh and Konijnenburg-van Cittert, 1989; Kujau et al., 2013), become a common floral element. The increase in humidity is well expressed by a lower xerophyte/hygrophyte ratio (avg. 60% xerophytes in core E-1). Climate Phase B is further characterised by increased abundances of *Cerebropollenites*-producing taxodiacean conifers, considered to belong to pioneer vegetations (Abbink et al., 2004). Contemporaneous strata from the eastern Netherlands are also marked by a strong decline in cheirolepidiaceans
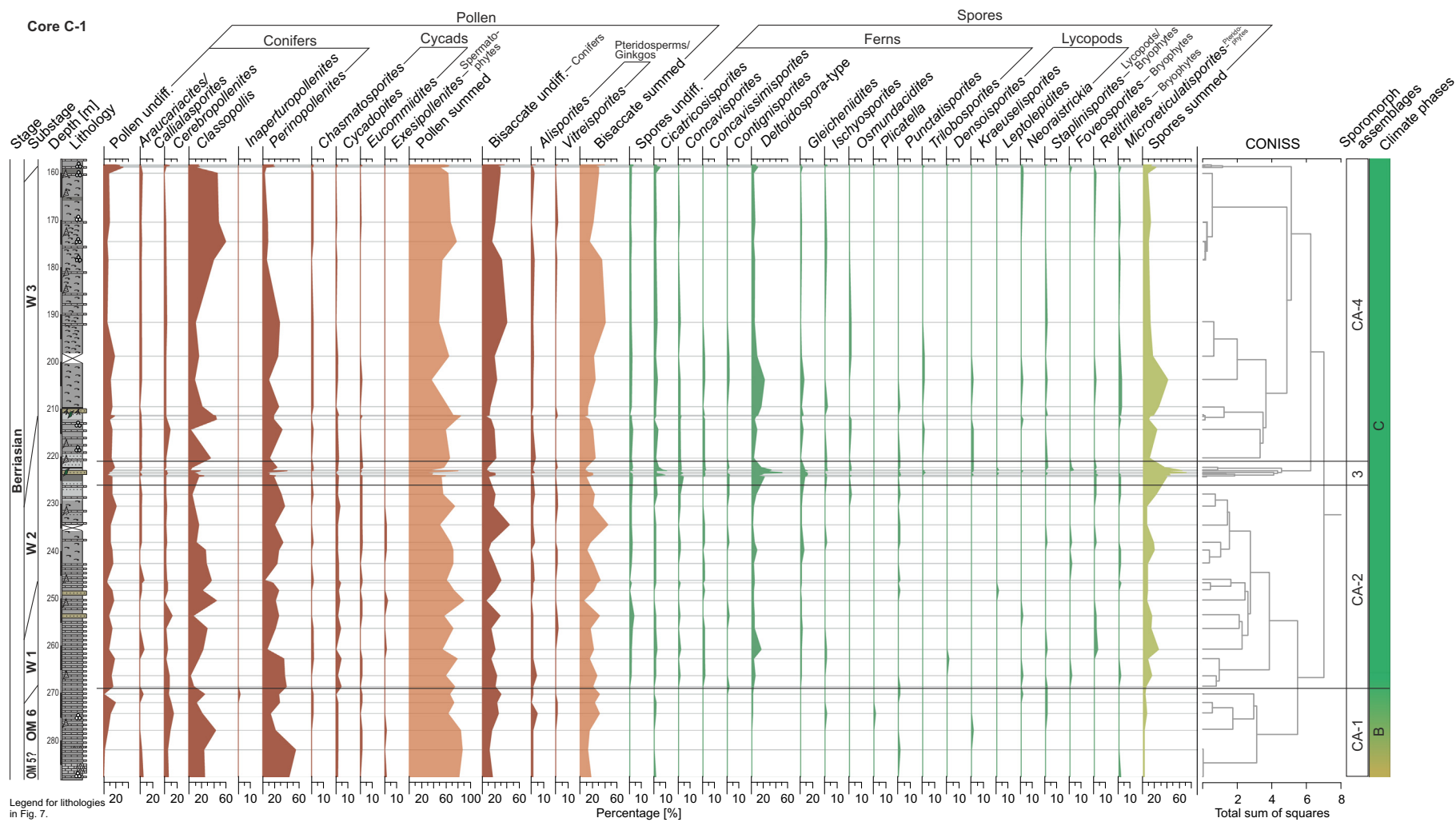
**Fig. 4.** Relative spore and pollen abundances (%) in core C-1 plotted against stratigraphy, depth (m) and lithology. The sporomorph assemblages are based on CONISS cluster analysis. Biostratigraphy is following Elstner and Mutterlose (1996). Climate phases are referring to humidity with orange colour representing low, green colour high moisture availability.

| Spore-pollen taxa or category | Botanical affinity |
|---|---|
| *Foveosporites, Retitriletes* | Bryophytes (Lycopodiaceae) |
| *Aequitriradites* | Bryophytes |
| *Staplinisporites* | Bryophytes and lycopods |
| *Leptolepidites* | Lycopods (Lycopodiales) |
| *Neoraistrickia, Densoisporites, Kraeuselisporites* | Lycopods (Selaginellaceae) |
| *Osmundacidites, Punctatisporites* | Ferns (Osmundaceae) |
| *Cyathidites/Deltoidospora/Dictyophyllidites, Concavisporites, Concavissimisporites, Trilobosporites* | Ferns (Dicksoniaceae, Cyatheaceae, Dipteridaceae, Matoniaceae) |
| *Contignisporites* | Ferns (Pteridaceae) |
| *Gleicheniidites* | Ferns (Gleicheniaceae) |
| *Cicatricosisporites, Ischyosporites, Plicatella, Varirugosisporites* | Ferns (Schizaeaceae) |
| Spores, undiff. | Mainly ferns |
| *Pilosisporites, Microreticulatisporites* | Pteridophytes |
| *Vitreisporites, Alisporites* | Pteridosperms/Ginkgos |
| *Chasmatosporites, Cycadopites, Eucommiidites* | Cycads |
| Bisaccate, undiff. | Conifers |
| *Araucariacites, Callialasporites* | Conifers (Araucariaceae) |
| *Cerebropollenites, Perinopollenites* | Conifers (Taxodiaceae) |
| *Classopollis* | Conifers (Cheirolepidiaceae) |
| *Inaperturopollenites* | Conifers (Cupressaceae) |
| *Exesipollenites* | Spermatophytes (Bennettitales) |

(Burger, 1966; base of pollen zone V) and an increase in *Cerebropollenites*-producing taxodiaceans (pollen zones U⁴ and V).

In contrast to the Tithonian assemblages, where *Botryococcus* is virtually absent (Heunisch and Luppold, 2015), climate Phase B is marked by relatively high abundances of this green algae accounting for up to 40% of the total particulate organic matter fraction (Fig. 6, core E-1). Blooms of *Botryococcus* have been recorded from shallow, partly saline lakes located in regions which are generally arid but subjected to occasional heavy rain fall (references in Tyson, 1995). Supported by sedimentological evidence as the occasional occurrence of evaporites, it is suggested that these Purbeck-type sediments were deposited in a semi-arid climate with temporary humid phases. Salinity fluctuations



**Fig. 5.** Diagram showing the relationship between the average percentage composition of terrestrial palynomorphs (spores, bisaccate pollen, and other pollen) and grain size of the samples from which they were extracted. A total of 58 samples was analysed from the interval 287.35–209.00 m of core C-1, representing all samples with a statistically significant amount of sporomorphs.

in the upper OM 5 and OM 6 have been documented from the Deister-Hils Embayment based on ostracod associations (Arp and Mennerich, 2008). Schnyder et al. (2009) in Dorset interpreted blooms of *Botryococcus*-type algae having been formed during the transitional climatic phase of early–mid Berriasian age under mainly freshwater to brackish, and locally hypersaline conditions.

### 5.4. Climate Phase C (Wealden 1–4)

Climate Phase C includes most of EA-3 and EA-4 as well as CA-2–4 (German Wealden). The pronounced increase in abundance of spores at the base of this interval suggests a spread of pteridophytes, particularly of ferns, under warm and humid conditions. This is supported by a further decrease of the xerophyte/hygrophyte ratio. Bryophytes and lycopods account for only small amounts of spores (up to 5%). Bryophytes usually grow under relatively humid conditions, but some groups can withstand long periods of drought (e.g., Abbink et al., 2004). In our data, small amounts of bryophyte spores occur in all three climate phases. The majority of extant lycopods occurs today in tropical regions, but many species are also common in temperate regions. Since lycopod spores occur regularly only in climate Phase C, they may be attributed to moist lowland environments as reported by other studies (e.g., Abbink et al., 2004).

Considerably high fluvial run-off from the hinterland is documented by deltaic deposits in the Deister-Hils Embayment. A well-preserved macroflora from an anastomosing river system of the upper deltaic plain is composed of a mixed ginkgoalean-conifer forest with abundant ferns (Pelzer, 1998; Pelzer et al., 1992; Pott et al., 2014). Horsetails presumably grew around small lagoons and in semi-inundated areas of the delta. Plants like lycopods, pteridosperms, Nilssoniales (a gymnosperm taxon different to cycads) and Bennettitales represent minor components of the vegetation. Pott et al. (2014) regarded this flora as intermediate, incorporating elements from both the English Wealden floras and more eastern (Siberian) floras.

In contrast to the fluvial deposits of the upper deltaic plain, in which only low amounts of *Classopollis* have been observed (Pelzer, 1998), the composition further "downstream" suggests a coastal swamp habitat subjected to brackish influences. Cheirolepidiaceans may have been either growing on sandy bars and coastal islands or even in mangrove-type communities. The latter was also suggested for the Wealden delta complex in south-eastern England (Hughes and Moody-Stuart, 1967; Batten, 1974). An enrichment of *Classopollis* in fine-grained distal strata, as documented in core C-1, was also registered by Hughes (1976).

In core C-1 high abundances of spores (up to 70%) have been registered in the Wealden 2–3 boundary deposits, including two thin siltstone bodies, a root horizon, and coal seams. These sediments clearly indicate a deposition close to the palaeo-shoreline. The overlying claystones were deposited in a more offshore position, documented by a major transgression (TE 1) at the base of Wealden 3 (Schneider et al., 2018). Consequently, fewer spores were deposited in these fine-grained strata which are enriched in *Classopollis* pollen and dinoflagellate cysts (Fig. 6). The lower-deltaic deposits from the upper part of core E-1, in contrast, yield very high amounts of spores.

In the course of the climatic change an enhanced freshwater input caused a reduction of water salinity. The brackish to evaporitic conditions of Phase A (OM 4–5) were succeeded by freshwater to brackish water bodies in the later part of Phase B and C. The presence of *Botryococcus*, prasinophytes, and the pioneering cyanobacterial species *Celyphus rallus*, which was adapted to freshwater–brackish transitional environments (Batten and van Geel, 1985), support humid conditions. Higher fluvial wash-in of soil nutrients and an incomplete mixing within the water column, resulting in the establishment of a salinity-driven stratification and reducing bottom water conditions, are associated with an organic matter enrichment in Phase B and C (Schneider et al., in work). This freshwater to brackish basin experienced several
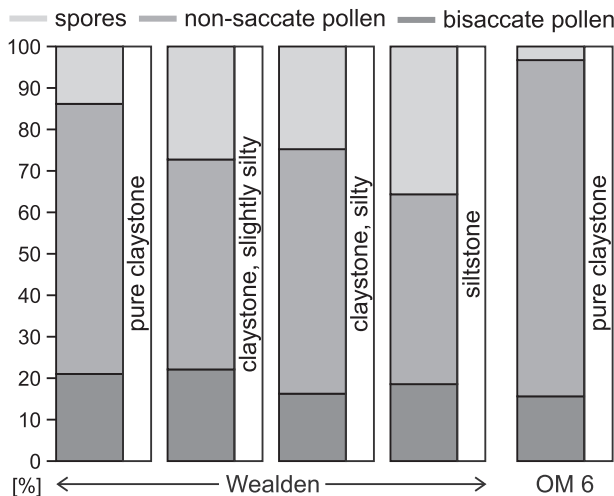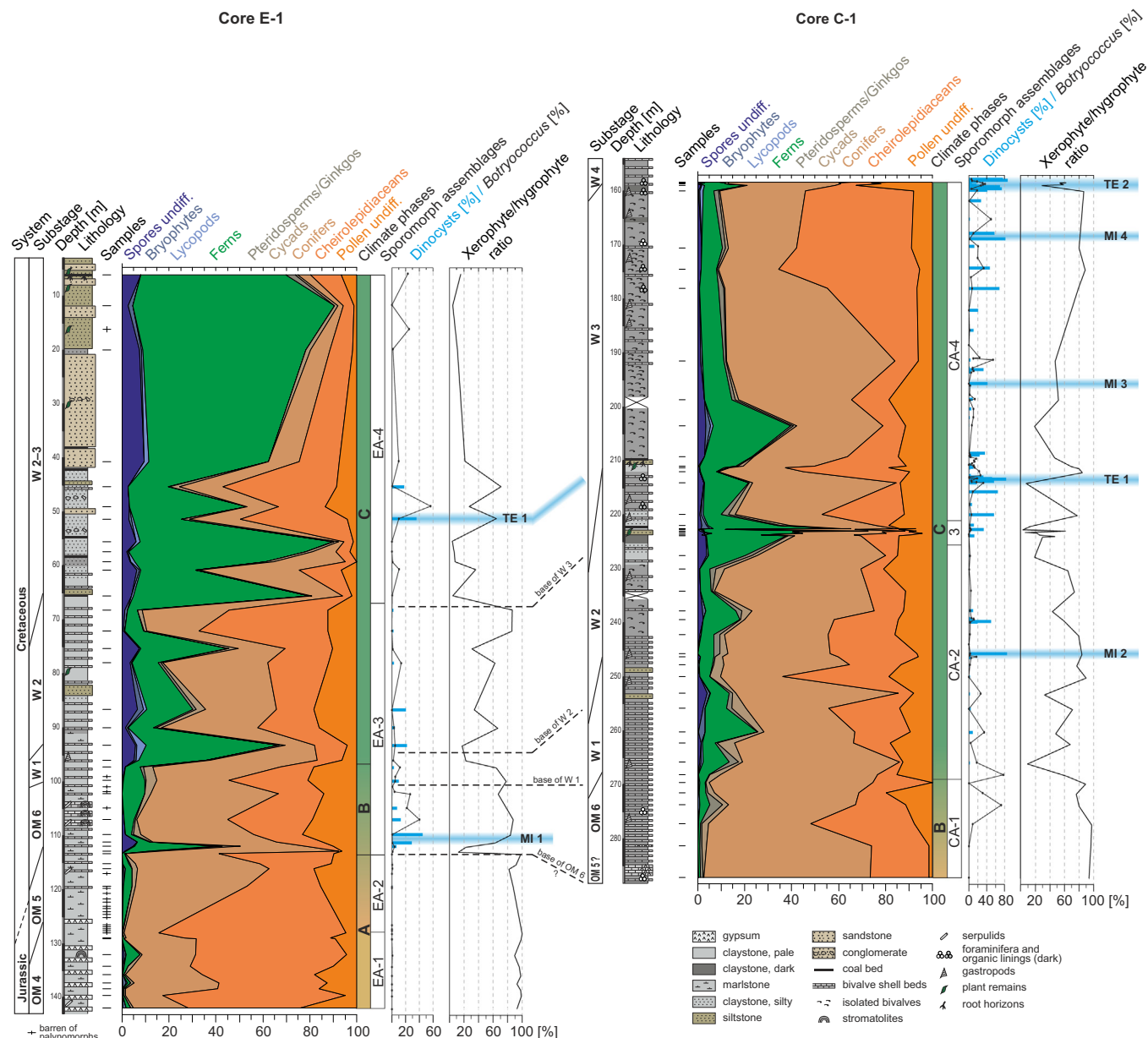
**Fig. 6.** Suggested reconstruction of latest Jurassic–earliest Cretaceous vegetation for cores E-1 and C-1. Seven principal floral groups have been recognized: cheirolepidiacean conifers, other conifers, cycads, pteridosperms/ginkgos, ferns, lycopods, and bryophytes. The extinct conifer family Cheirolepidiaceae is emphasized here due to its climatic importance. The abundances of aquatic palynomorphs are given in percent of total organic particles. Dinoflagellate cyst and *Botryococcus* abundances as well as marine transgressive events (TE) and marine incursions (MI) are adopted from Schneider et al. (2018).

short-lived episodes of marine flooding, entering the basin presumably from the west via the Netherlands as documented by dinoflagellate cyst assemblages and foraminifera (Strauss et al., 1993; Schneider et al., 2018).

## 6. Correlation with other northwest European J–K boundary successions

The palynological data from the LSB suggest a three-phase climatic trend which can be embedded in the high-resolution stratigraphic framework based on palynology (spores, pollen, dinoflagellate cysts) and ostracods (Schneider et al., 2018). The late Tithonian–earliest Berriasian arid interval (Phase A) is followed by an early–mid Berriasian transitional phase (Phase B) and a mid–late Berriasian humid period (Phase C). In the type-section of the Purbeck in England a synchronous tripartite climatic evolution has been documented based on clay

mineralogy (Schnyder et al., 2006) and biomarkers (Riboulleau et al., 2007). Purbeckian strata including the Soft Cockle Beds were deposited in an evaporitic environment (equivalent to Phase A), followed by freshwater, brackish to locally evaporitic conditions during the deposition of the Cherty Freshwater Beds and Cinder Bed (equivalent to Phase B). The peak in spore abundances at the base of Phase B in three samples of core E-1 (12–40% spores) may indicate a pronounced short-lived humid phase. A short-term rise in humidity has also been observed in the middle Cherty Freshwater Beds from England (Hunt, 1985), documented by an increase in spore diversity and abundance and a decline of *Classopollis* (Fig. 2). Palynological data from Schneider et al. (2018) support a correlation of these two horizons, being closely associated with a marine incursion (MI 1) at both localities.

Above the Cinder Bed, freshwater to brackish conditions prevailed and a rise in kaolinite in the overlying Intermarine Beds points to an increase in humidity (Schnyder et al., 2006). This level corresponds to

Phase C in the LSB. The former xerophytic coniferalean vegetation in northwest Europe was replaced by a spore-dominated flora with a large diversity of ferns, whilst temperatures remained relatively warm (e.g., Hallam, 1984; Abbink, 1998; McArthur et al., 2007). Although *Classopollis* is less common in Phase C, cheirolepidiaceans were still an important component of the coastal floras.

This work provides 1. additional information about the climatic change across the J–K boundary interval, 2. new quantitative palynological data that is biostratigraphically well dated. Correlation of the non-marine sequence in northern Germany with the terrestrial but near-marine depositional setting in southern Sweden and the marine succession of the southern North Sea is possible via the type-section of the Purbeck in Dorset (Fig. 2; Schneider et al., 2018). Accordingly, the onset of this climatic change occurred in the upper part of the Boreal *Subcraspedites lamplughi* ammonite Zone, thereby slightly earlier than documented by sporomorph assemblages from the southern North Sea (Abbink et al., 2001). The end of the transitional phase and the permanent establishment of humid conditions occurred synchronously in the middle *Heteroceras kochi* ammonite Zone.

The change from a dry climatic phase to a more humid one, recorded in northwestern Europe and along the northern margin of the Tethys is also recorded from Tunisia. The results of Schnyder et al. (2005) based on clay mineral assemblages document that the climatic shift occurred synchronously on both margins of the Tethys. In contrast, Lindström and Erlström (2011) documented from the Vomb Trough (southern Sweden) a climatic change from seasonally dry to constantly humid conditions already for the Tithonian–Berriasian boundary interval. This record pre-dates the shifts reported from England, Germany, and the southern North Sea by ca. 2–3 million years. The current study improves the stratigraphic framework and the palaeogeographic setting for the climate change of the J–K boundary interval in northern Europe. Our findings will stimulate new studies and improve the understanding of which factors caused this large-scale change.

## 7. Conclusions

Palynofloral and sedimentological data from the Lower Saxony Basin document a progressive climatic trend, which can be correlated with a synchronous climate shift recorded in the English Purbeck. Correlation with strata from the southern North Sea clearly indicates that the climatic change began in the upper part of the Boreal *Subcraspedites lamplughi* ammonite Zone and culminated in the middle *Heteroceras kochi* Zone. This suggests a 2.5-million-year long period of increasing humidity. Following the palynological data three climatic phases (A–C) can be differentiated.

Phase A: In the Tithonian–earliest Berriasian arid conditions prevailed, favouring the growth of coastal forest communities with dominant *Classopollis*-type drought-resistant conifers. The evaporitic sequences of Phase A, which are accompanied by stromatolites and serpulids, are barren of the green algae *Botryococcus*.

Phase B: The early–mid Berriasian shows an increasing diversification of spore-producing floras at the top of OM 5 (former Katzberg Member). A decline in cheirolepidiaceans was accompanied by the spread of scarce pioneer vegetation, comprising *Cerebropollenites*-producing taxodiaceans, cycads, pteridosperms and ferns. Deposition took place in an overall brackish to evaporitic environment with fluctuating salinities, as indicated by the episodic precipitation of evaporites and changing ostracod associations. These conditions together with enhanced fluvial wash-in of soil nutrients induced blooms of the green algae *Botryococcus*. This long-term trend of Phase B is punctuated by a pronounced short-lived humid phase in the lower OM 6 (former Serpulit Member) which correlates with a short-term rise in humidity reported from southern England.

Phase C: The mid–late Berriasian is marked by an increase in spore diversity and abundance, documenting the proliferation of diverse fern associations in fluvio-deltaic mixed swamp forests under warm, humid conditions. Cheirolepidiaceans remained an important component of the coastal flora possibly belonging to mangrove-type communities. The change in the composition of palynomorphs is recorded at the base of the German Wealden. Brackish to evaporitic conditions are replaced by a brackish to freshwater lake system with abundant dinoflagellate cysts and *Botryococcus* algae. Although palynological and sedimentological data from both cores document humid conditions for this phase, differences in the sporomorph compositions especially in the upper part are clearly influenced by hydrodynamic sorting.

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